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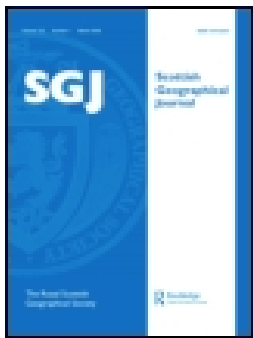
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The pre-glacial landscape of Antarctica

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ABSTRACT

The geomorphology of the hidden subglacial landscape of Antarctica is relevant to our understanding of the stability of the Antarctic Ice Sheet and also to that of global interactions between plate tectonics and surface processes. We believe that geomorphology has much to contribute, but that the lack of coherent hypotheses about the origins of the subglacial landscape is holding back understanding. This paper approaches the problem by using southern hemisphere land masses in Africa and Madagascar as analogues. We find that the Antarctic landscape evolved in a similar way to passive margin evolution in southern Africa. Rifting associated with the breakup of Gondwana changed river base levels and caused rapid erosion on the flanks of rifts and was accompanied by the uplift of rift-margin mountains. Rift-margin plains, often coastal or extending inland along large rivers, are backed by an escarpment, while low-gradient continental river basins characterised the interior of Antarctica. In East Antarctica ice has removed pre-existing regolith from lowlands and excavated 2–3 km troughs below sea level along the course of major trunk rivers. The micro-continents of West Antarctica are comparable to Madagascar and apparently share a similar topography with coastal plains, backing escarpments and interior plateaux.

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Introduction

The aim of this paper is to characterise the hidden landscape of Antarctica before it was covered by ice sheets. We believe that the geomorphology of the current landscape can provide significant insights, but that the lack of coherent hypotheses about the fluvial origins of the subglacial landscape is holding back understanding. We approach this problem by using southern land masses such as Africa and Madagascar as analogies of similar landscapes that have escaped glaciation. Such continents and micro continents were, together with Antarctica, part of Gondwana and experienced a similar story of landscape evolution during continental breakup.

It is remarkable to reflect that the subglacial landscape of Antarctica is less well known than that of the Moon and Mars and yet it is important for our understanding of how planet Earth works. A key concern is the stability of the Antarctic ice sheet in a

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warming world and its effect on global sea level (Joughin & Alley, 2011; Scherer, DeConto, Pollard, & Alley, 2015). Major outlet glaciers draining significant parts of the ice sheet are deepening their beds below sea level due to glacial erosion and are thus increasingly susceptible to melting by warm ocean waters. As beds deepen, the gateways through which ice flows get larger, meaning more ice can drain from the ice sheet. How close are such glaciers to a threshold of instability? In order to answer such a question, it is important to have knowledge of the trajectory of million-year landscape evolution and its role in pre-conditioning the Antarctic Ice Sheet to instability.

The theory of plate tectonics has transformed our approach to understanding geomorphology. The mid-twentieth century division into short-term studies of process and long-term reconstructions of landscape evolution has been replaced by an integrated subject that can bring important insights into the processes of plate tectonics and its two-way interaction with geomorphology (Bierman & Montgomery, 2014; Bishop, 2007; Summerfield, 1981). Observations on the ground help tune and test geophysical models of the working of the crust (Summerfield, 2000). The key insight is that spatial variations in surface erosion and deposition affect patterns of isostatic uplift and depression and this in turn influences the geophysical processes accompanying plate tectonics. Moreover, new dating techniques over a range of time scales allow rates of change to be established. Antarctica has contributed relatively little to this new geomorphology since studies have been restricted to a few scattered ice-free areas and there is uncertainty about the interpretation of subglacial landforms. One example of the latter concerns flat surfaces close to sea level backed by cliffs/escarpments which could represent either marine platforms or fluvial lowlands (Rose et al., 2015). Each explanation has different implications for understanding landscape evolution and the geophysical processes involved.

Understanding the Antarctic Ice Sheet depends on knowledge of the relationship between the pre-existing fluvial landscape and the way the ice sheet flows across and modifies the landscape. There has been steady progress in understanding the ice sheet and its evolution through radio-echo sounding of the bed (Siebert, Ross, & Le Brocq, 2016), the drilling of ice cores (EPICA, 2004), ice-sheet modelling (Jamieson & Sugden, 2008; Pollard & DeConto, 2009) and satellite remote sensing of the surface (Rignot, Mouginot, Morlighem, Seroussi, & Scheuchl, 2014). In comparison, there has been less progress in interpreting the bed. There are studies of mountains exposed around the continental margins (Baroni, Noti, Ciccacci, Righini, & Salvatore, 2005; Näslund, 2001; Sugden, Denton, & Marchant, 1995, 2017), the subglacial Gamburtsev Subglacial Mountains (Rose et al., 2013), some predictive ice-sheet modelling based on glaciological principles (Jamieson, Sugden, & Hulton, 2010), and some broad interpretations of the degree to which the subglacial landscape has been modified by ice (Jamieson et al., 2014). Given this slim background, we believe that a study of fluvial analogues from other southern continents will add perspective and improve our understanding of the subglacial landscape of Antarctica.

The Scottish Geographical Journal is a good place for such an article. Scotland has a strong tradition in physical geography and geomorphology and a long-standing interest in Antarctica. After all the Royal Scottish Geographical Society (RSGS) supported the Scottish National Antarctic Expedition of W.S. Bruce in 1902–1904 and E.H. Shackleton was secretary of the RSGS in 1904–1905. Further, the potential for geomorphology in Antarctica is already there in openly available data sets such as BedMap 2 (Fretwell et al., 2013)

and will increase as new higher resolution data sets become available. A well-posed interrogation of existing data sets has already produced new insights into subglacial Antarctic volcanoes (Van Wyk de Vries, Bingham, & Hein, 2017). We hope that our approach to Antarctica and analogous southern continents will generate many more such questions.

Antarctic background

The Antarctic Ice Sheet measures some 4500 km across and the ice rises to an elevation in excess of 4200 m in East Antarctica and 3000 m in West Antarctica. It is dome shaped with steeper slopes nearest the periphery. Much of the periphery is drained by outlet glaciers focussed on coastal embayments or sometimes breaching high mountain rims. The volume of ice in the ice sheet is 27 million km³, equivalent to a potential contribution to sea level of 58 m (Fretwell et al., 2013). Lying between East and West Antarctica are the large coastal embayments of the Weddell Sea and Ross Sea. Here are floating ice shelves, hundreds of km across, with flat surfaces and ice over 1000 m in thickness at the grounding line and 100–200 m on the seaward flank.

The subglacial landscape is shown in relation to present sea level in Figure 1; it has been isostatically rebounded to account for removal of the modern ice load. In the centre of East Antarctica and inland of the lowlands of the Lambert Graben are the Gamburtsev Subglacial

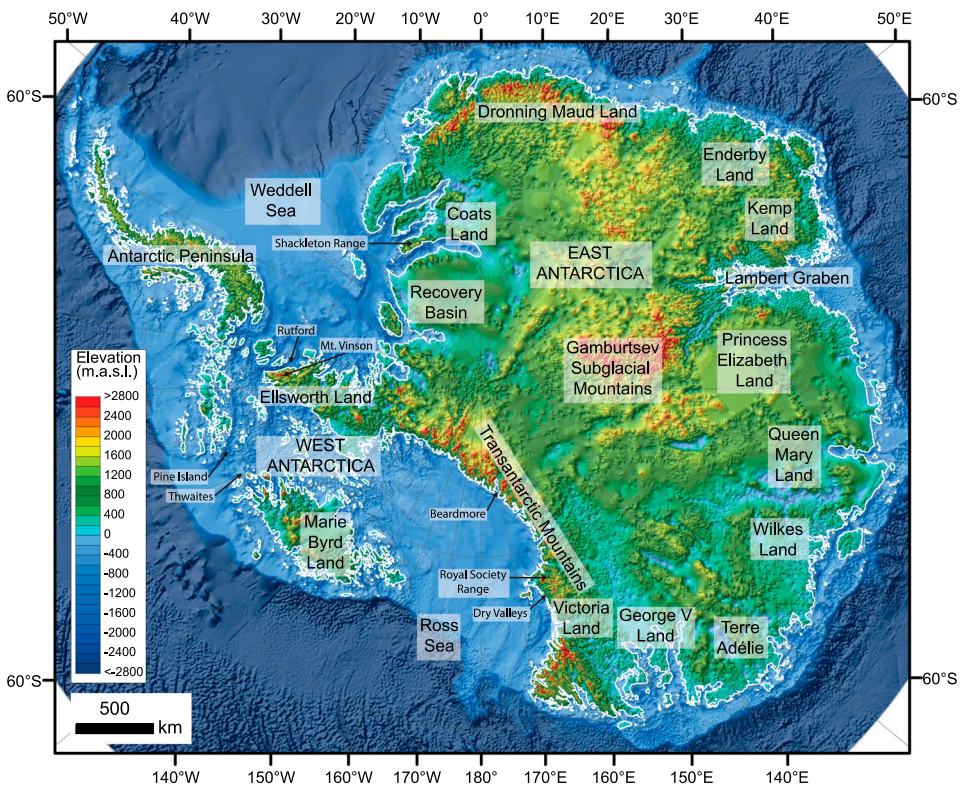


Figure 1. The bedmap2 bed topography of Antarctica (Fretwell et al., 2013) rebounded after the removal of present ice load (Jamieson et al., 2014). The white line indicates the coastline under these rebounded conditions.

Mountains 3400 m high. There are also peripheral mountains with elevations in excess of 3000 m and in the case of Dronning Maud Land and the Transantarctic Mountains they slope down gently on the inland flank to lowlands. In Dronning Maud Land a coastal lowland lies between the mountain front and coast while the Transantarctic Mountain front overlooks the marine basin of the Ross Sea. In many locations, wide troughs extend several hundred km towards the coast and are below sea level. In West Antarctica, three main mountain blocks are separated by marine basins up to 2555 m deep. Ellsworth Land includes the highest mountain in Antarctica (Mt. Vinson, 4892 m); the others are the mountain blocks of Marie Byrd Land and the Antarctic Peninsula with elevations over 2000 m.

Plate tectonic history helps explain the major landscape features and shows the links with the southern continents (Figure 2). East Antarctica comprises a coherent and ancient craton, similar to those in Africa, Australia, India, Madagascar and South America. They were all part of Gondwana before separation that began 180–160 million years ago (Dalziel, 1997). South America and Australia were finally separated from Antarctica by open seaways from Antarctica around 35–40 million years ago. Remarkably, the continuation of the Lambert Graben can be identified in eastern India (Harrowfield, Holdgate, Wilson, & McLoughlin, 2005), while the basalts and dolerite sills caused by volcanic activity associated with break up are found in all southern continents. In West Antarctica stretching of the crust has produced a series of micro-continents surrounded by thinner crust, resembling a series of horst and graben structures. The Ellsworth and Marie Byrd Land mountain blocks have separated from the East Antarctic craton, while the Antarctic Peninsula block is derived from the Andes in South America (König & Jokat, 2006). The West Antarctic Rift System has involved stretching of 300–600 km beneath the Ross Sea and extends towards the Antarctic Peninsula and Weddell Sea (Wilson et al., 2012). Volcanoes still mark the rift system, including 3794



Figure 2. Plate configuration of Gondwana at the early stages of breakup about 120 million years ago. After Lawver, Gahagan, and Coffin (1992).

m-high volcanic edifices such as Mount Erebus that have been active for the last 19 million years and subglacial volcanoes, some of which have erupted in the last few thousand years (Iverson et al., 2017).

The climate over the last tens of millions of years saw Antarctica and indeed the rest of the globe move from a greenhouse to an icehouse world. The oxygen isotope curve derived from deep sea sediments and shown in Figure 3 is a combined measure of the volume of ice on land and ocean temperature over the last 40 million years (Miller et al., 2005). The panels show the likely Antarctic ice sheet extents associated with the evolving global environment. Ice sheet growth first occurred some 34 million years ago during cooling at the Eocene-Oligocene transition, although there may have been mountain glaciers at an earlier stage. Continental-scale glaciation was triggered by changes in atmospheric CO₂ and the deepening of seaways between Antarctica and the southern continents of Australia and South America (DeConto & Pollard, 2003; Kennett, 1977). There followed a period of nearly 20 million years during which Antarctic ice sheets waxed and waned in response to Croll-Milankovitch cycles. Deposits at Cape Roberts near the Dry Valleys reveal at least 33 cycles of advance and retreat, with meltwater abundant (Naish, Woolfe, & Barrett, 2001). At about 15 million years ago there was a further fall in air and sea temperatures. This caused the ice sheet to extend to its maximum over the continental shelf, perhaps several times, before withdrawing and stabilising at the present coastline. Since that time, the extent to which the ice margin has retreated and advanced is uncertain. For example, there is much debate about how much ice was lost from the Antarctic ice sheet during warm periods such as that of the Pliocene ~3–5 million years ago (Hein et al., 2015; Scherer et al., 2015). Further cooling since the Pliocene saw the world enter the era of large Pleistocene ice sheets in the northern hemisphere. The growth of northern ice sheets lowered global sea level and this permitted coastal expansion in Antarctica (Bentley et al., 2014).

The changes in climate are reflected in changes in vegetation. In the early Eocene some 50 million years ago Antarctica was warm and wet, at least near the coast. The Antarctic Peninsula supported forests similar to the low altitude Valdivian rainforests in southern

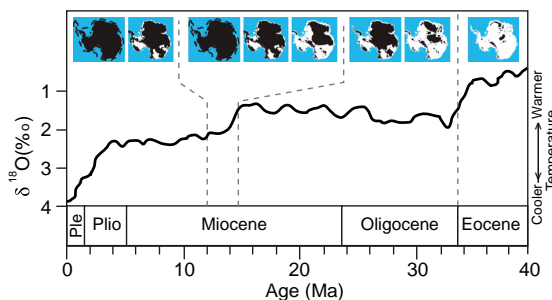


Figure 3. The greenhouse-icehouse transition in Antarctica. The oxygen isotope curve (black) is a measure of world temperature combining ice volumes and ocean temperatures and comes from analysis of deep-sea sediments (Miller, Wright, & Browning, 2005). The ice extents for different stages (shaded black) come from ice sheet modelling (Jamieson et al., 2010). The Oligocene to middle Miocene saw ~ 33 glacial cycles with meltwater abundant, probably similar to conditions in present-day Greenland. Subsequent Miocene cooling saw the Antarctic Ice Sheet advance to the edge of the continental edge, perhaps on several occasions.

Chile flourishing in a frost-free environment (Francis et al., 2009). The forests were dominated by southern beech (*Nothofagus*), conifers, ferns and horsetails. Parts of the East Antarctic coast are thought to have been warm enough to support near-tropical forests containing palm trees, the presence of the latter implying that winter temperatures must have remained above freezing. Subsequent cooling saw tropical vegetation giving way to temperate rainforest (Pross et al., 2012). Antarctic warmth reflected the north–south exchange of ocean water that existed prior to the development of the Antarctic circumpolar current. By the time of ice-sheet glaciation at ~34 Ma the warmth-loving plants had been replaced by cooler vegetation comprising several species of shrubby southern beech, mosses and a few ferns. Tundra vegetation survived in places until around 15 million years ago. Clay minerals mirror the vegetation changes. Weathering in the warm and continual wet conditions of the early Eocene favoured the production of smectite, while the subsequent change to cooler, seasonal, wet conditions favoured illite (Dingle & Lavelle, 1998; Ehrmann, Setti, & Marinoni, 2005).

Fluvial action on passive continental margins

Given the tectonic and climatic background above the next step is to discuss how rivers are likely to have responded to the break-up of Gondwana and to discuss the types of landform that result. This is the realm of rifts, faults and passive continental margins. There are two major effects of rifting (Gilchrist & Summerfield, 1990; Kooi & Beaumont, 1994). First, rivers flowing into the faulted rift are rejuvenated by a drop in base level and begin rapidly to erode the landscape to the new base level. This is because the steeper river gradients and valley slopes lead to accelerated rates of erosion. Second, extension (pulling apart) of the rift leads to further faulting, crustal thinning and thermal uplift of the rift flank often sufficient to disrupt existing river drainage. As erosion proceeds, the isostatic response to sedimentary loading in the rift and unloading of the rift flank due to erosion causes local isostatic subsidence and uplift. The stiffness of the crust means that the location of the axis of maximum marginal uplift may be several hundred km distant. Numerical models and observations suggest that the typical landforms resulting are a shallow gradient plain cut by river action adjacent to the rift (or sea) and that the plain is commonly backed by an escarpment (Gilchrist & Summerfield, 1990; Kooi & Beaumont, 1994). In places, rivers from the interior may have maintained valleys through the escarpment and contribute a more complex geomorphology. Under such circumstances, the effect of the change of base level may spread laterally from the main river channel in the interior. Rates of landscape erosion and deposition are highest in the tens of millions of years following initiation of the change in base level (Fitzgerald, 1992).

A classic example of the evolution of such a passive continental margin is the case of Namibia in southern Africa (Figure 4) (Cockburn, Brown, Summerfield, & Seidl, 2000). Here is an escarpment with a relief of 1000 m and an elevation of 2000 m situated 170 km from the present coastline and close to the main drainage divide. Between the escarpment foot and the coast is plain with a regional slope of 0.3° marked by river channels sourced mainly from the escarpment. Figure 4 shows the amount of rock removed by erosion at two intervals. Between the time of rifting 130 million years ago and the start of the Oligocene

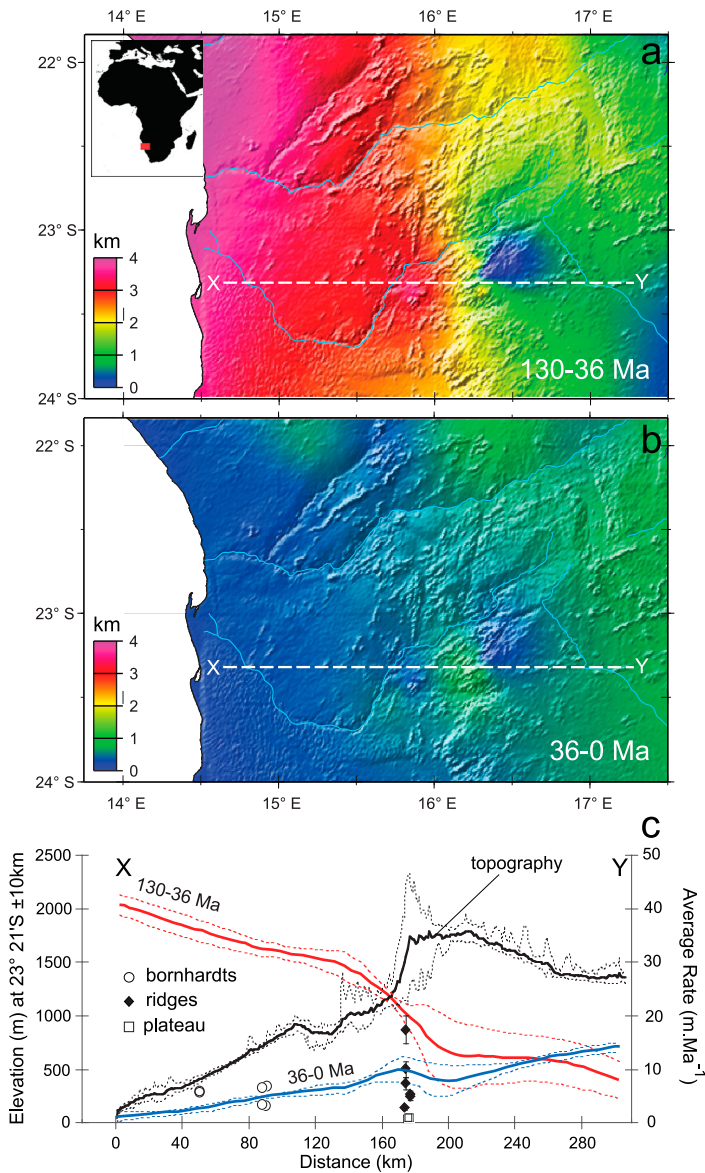


Figure 4. Denudation associated with passive margin evolution in Namibia, southern Africa (from Cockburn et al., 2000). (A) Post-break-up (130 Ma) to the end of the Eocene (36 Ma). (B) Post-Eocene. (C) Variations in mean denudation rates (solid coloured lines) and the standard deviation (dashed coloured lines) across the margin for each of the two periods. Mean (solid black line) and maximum and minimum (dashed black lines) topography of the region is from GTOPO30 digital elevation data. Denudation rates (symbols) modelled for the cosmogenic isotope data are the mean values of paired ^{10}Be and ^{26}Al measurements.

36 million years ago over 4 km has been eroded from the coast and the thickness decreases inland across the plain. Mean denudation rates for the coastal plain averaged 40 m per million years but are likely to have been highest after initial rifting. In the last 36 million

years erosion of the plain has occurred at a mean rate of 5 m per million years. Inland of the escarpment rates of denudation since break-up have remained constant at around 10 m per million years. The significance of this example is that it supports the view that most erosion takes place soon after rifting when river channels and valley slopes are steepest, and that the landscape may be denuded of many km of rock near the rift margin. Moreover, one can expect the formation of coastal or rift margin plains with shallow but distinct regional gradients, and often significant escarpments inland and parallel to the rift or major rivers.

Continental interiors tend to have low-relief landscapes with large alluvial rivers draining large basins. Thick weathering profiles on gentle slopes with rounded corestones set in a soft yellow-red regolith are characteristic of low-relief landscapes in warm humid areas and are well illustrated by photographs from Madagascar (Figure 5). In an environment dominated by weathering, quartz-rich rocks are resistant and typically form upstanding bornhardts/inselbergs rising hundreds of m above the surrounding plain (Figure 5). Many inselbergs display sheeting parallel to the surface. In more arid continental interiors surface wash is the dominant process and bare rock surfaces with straight slopes are typical. An example of the latter would be the escarpments and inselbergs of Monument Valley in Arizona.



Figure 5. Examples of weathered landscapes as a result of warm, humid climate similar to Antarctica during the Eocene. (A and B) Corestones set in soft, deeply weathered regolith near Antananarivo, Madagascar. (C) As regolith is removed, corestones begin to armour the landscape, south central Madagascar. (D) A bornhardt/inselberg with sheeting parallel to the surface emerges following the removal of the surrounding weathered material, south central Madagascar.

Fluvial passive margin landscapes in Antarctica

Studies of landscape evolution in the McMurdo Dry Valleys that comprise part of the Transantarctic Mountains suggest that at least in this ice-free region of Antarctica the fluvial signature is strong and tells of classic passive margin evolution. The fluvial nature of the Dry Valleys was first noted by some geologists and geographers accompanying the expeditions of Scott and Shackleton early in the twentieth century (Priestley, 1909). The Dry Valleys themselves are sinuous with a dendritic pattern and some originate at the crest of the mountain divide. In the Dry Valleys mountain block there are two escarpments with a relief of over 1000 m, each at the back of flat surfaces bordering the Ross Sea, the latter a zone of thinner, stretched crust (Figure 6). Normal faults occur at the rift margin and offshore is a basin containing sediment over 8 km thick. The seaward escarpment coincides with a fault while the inland escarpment ~100 km from the coast, is erosional in origin (Figure 6). Beacon sandstone sediments punctuated by dolerite sills lie above a granite basement that rises in elevation towards the coast at the rift margin. Resistant dolerite sills form intermediate surfaces. Much work has been carried out in the region and is summarised elsewhere (Kerr et al., 2000; Sugden & Denton, 2004). Figure 6 shows the stages of landscape evolution, with an idealised immediate post-rift start, the subsequent erosion of 4–5 km of rock at the coast, and the isostasy that caused uplift of the basement and further faulting at the coast. Dating shows that most of the erosion and associated uplift occurred around 55 million years ago (Fitzgerald, 1992), indicating a phase of rapid fluvial erosion in the Eocene, probably due to a change in base level related to crustal extension in the Ross Sea.

The present morphology of the Transantarctic Mountains varies from block to block. In Victoria Land, the mountains are highly dissected by valley glaciers diverging somewhat radially from the mountain crest. Quantitative analysis shows the valley pattern is inherited from an earlier fluvial network (Baroni et al., 2005). South of the Dry Valleys is the Royal Society Range with an escarpment over 4000 m in elevation within 25 km of the coast. Immediately north of the Dry Valleys and separated by a formerly sinuous valley now occupied by Mackay Glacier is the Convoy Range with the main erosional escarpment ~50 km from the coast. The most likely cause of such differences is the varying relationship between the axis of maximum uplift due to crustal flexure and the location of the initial fluvial drainage divide (Gilchrist & Summerfield, 1990). Throughout its length, the uplifted plateau of the Transantarctic Mountains comprises discrete blocks separated by outlet glaciers that flow to the Ross Ice Shelf. Between latitudes 74 and 87°S, there are 32 such major glaciers. Some, such as Beardmore Glacier which is 15–45 km wide, have exploited a pre-existing river valley that was cut close to sea level (Webb, 1994). It seems fair to conclude that the uplifted plateau edge of the Transantarctic Mountains is a classic example of a rift margin bordering a zone of extended crust. Rivers originating at the crest have eroded a wedge of rock close to the coast. Other rivers, such as that carving Beardmore Valley, may have maintained a valley across the uplifting mountains. A notable feature throughout much of the length of the Transantarctic Mountains is the gentle slope of the inland flank of the mountains leading down towards the low-lying plains of the interior.

There are other features in the subglacial landscape of Antarctica that display similar passive margin landscapes. Perusal of Figure 1 reveals that in Dronning Maud Land is a

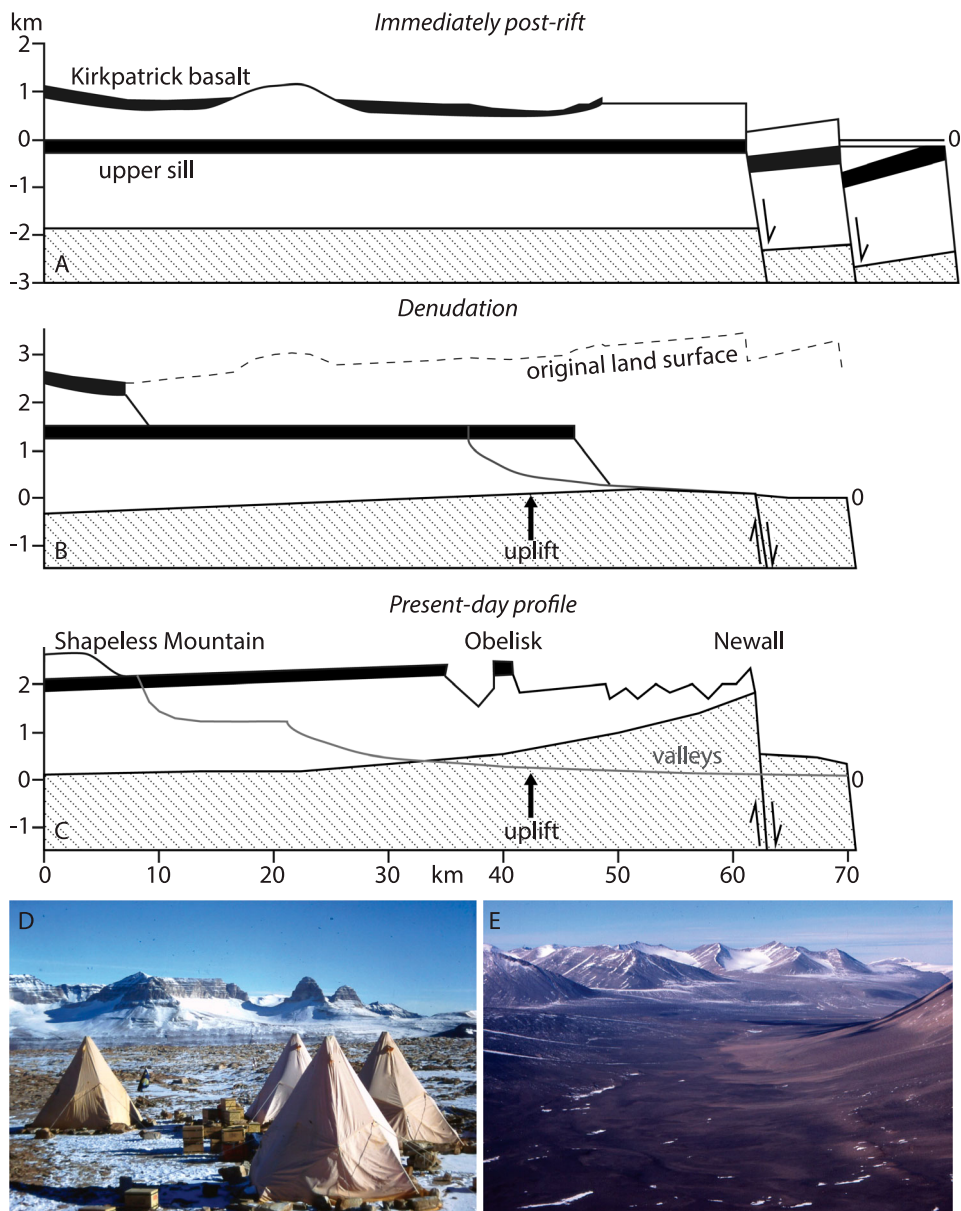


Figure 6. Schematic evolution of Dry Valleys (modified from Kerr, Sugden, & Summerfield, 2000). (A) Initial lithospheric extension and rifting. (B) Fluvial denudation to the new sea level removes ~4 km from the coast and is accompanied by isostatic uplift and faulting. (C) Rivers deepen the Dry Valleys and enhance crustal uplift that raises the basement rocks at the mountain front. (D) The main erosional escarpment near Shapeless Mountain seen from Sesrumnir Valley in the Asgard Range. The isolated mountains in front of the escarpment are part of the Olympus Range. The escarpment can be seen as the equivalent of the Drakensberg escarpment in South Africa. (E) View from Balham Valley towards Victoria Valley in the heart of the Dry Valleys, showing in the distance rectilinear slope elements.

coastal plain about 100 km wide backed on its inland margin by an escarpment and mountain crest comparable in elevation to the Transantarctic Mountains. Moreover, the interior flank also slopes down to a lowland plain, thus displaying the main features of a passive margin. In Enderby Land, Kemp Land and Princess Elizabeth Land there is a coastal plain 100–300 km wide backed by an escarpment. The interior flanks of the escarpment crest vary from coherent upland plateaux to dissected uplands. Lowland plains some hundred km across characterise the sides of the Lambert Graben. The 1000 km long graben formed some 250 million years ago and offshore sediments show it was drained by a river flowing to the coast (Thomson, Reiners, Hemming, & Gehrels, 2013). The plains adjacent to the river are backed by clear escarpments with a relief of 1000–2000 m. Some escarpments are cut into the northern flank of the Gamburtsev Subglacial Mountains. The mountains themselves, rising to elevations of 3400 m, show clear evidence of fluvial dissection with dendritic valley patterns and fluvial characteristics such as consistent stream ordering and concave valley long profiles since modified by ice (Figure 7) (Rose et al., 2013). In view of these fluvial characteristics, it seems reasonable to suppose that at an early stage low plains extended from the main river in the Lambert Graben eroding laterally into plateaux and mountains. Analysis of offshore sediments shows that initial rates of denudation were highest about 250 million years ago when the graben formed (Lisker, Brown, & Fabel, 2003). During the following hundred million years rates were low at 20 m per million years until glaciation began some 34 million years ago (Thomson et al., 2013). One implication is that most fluvial dissection of the Gamburtsev Subglacial Mountains must also be equally ancient. Other large interior

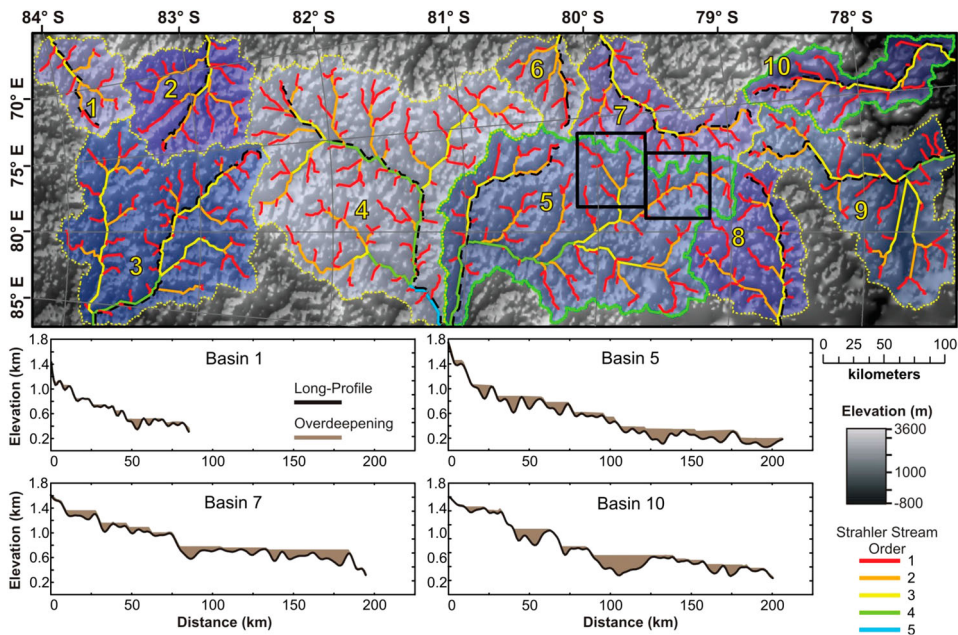


Figure 7. A fluvial landscape fingerprint in the subglacial Gamburtsev Subglacial Mountains (Rose et al., 2013). The map shows stream ordering and the dendritic pattern of valleys in individual basins. Below, the long profile of the main rivers displays a concave upwards profile with local overdeepened basins typical of glacial modification.

basins may also be marked by plains bounded by escarpments. Such a landscape has been recently reported from George V Land (Paxman et al., 2018) where radar measurements of the buried landscape reveal a series of elevated flat surfaces that are bounded by an escarpment that is currently 400–500 km inland of the modern coast. Escarpments related to river valleys have also been described in the Shackleton Range on the border of Recovery Basin (Paxman et al., 2017; Sugden et al., 2014).

Some perspective on these interior river basins can be obtained through hydrological modelling and comparison with other interior basins in the southern continents (Figure 8). With few exceptions, the modelling shows coherent dendritic patterns with 5 major river basins radiating from the Gamburtsev Subglacial Mountains. A group of basins drains the interior lowlands of Dronning Maud Land and Recovery Basin into the Weddell Sea while other large river basins flow into the Lambert Graben, Wilkes Land and George V Land, the latter draining the lowland inland of the Transantarctic Mountains. The basins are comparable in size to those in other southern continents. The Recovery, Lambert and Wilkes river basins have areas of 0.73×10^6 , 0.79×10^6 & 0.78×10^6 km² respectively. This compares with the river basins of the Orange (0.89×10^6 km²) and Murray (1.14×10^6 km²) (Summerfield & Hulton, 1994). In addition to the continental-scale drainage basins, a very large number of small drainage basins are found around the margin of Antarctica and reflect the position of the scarp that has evolved as a result of the breakup of Gondwana and the subsequent headward erosion of coastal rivers. These range in scale from 50 to 400 km in length. Summerfield and Hulton (1994) note, based on a study of the world's rivers (not including Antarctica), that the relief ratio (basin relief/basin length) of a drainage basin is a simple predictor of rates of denudation. Basin relief ratios for Antarctica are shown in Figure 8. As examples, the Recovery, Lambert and Wilkes basins have relief ratios of 0.0017, 0.0021, and 0.0015, respectively. These compare most closely to the Amazon (0.0017), Orange (0.0022) and Danube (0.0018) rivers (Summerfield & Hulton, 1994).

An alternative approach to understanding fluvial denudation was presented by Syvitski et al. (2003) in which they suggested that climate, basin area and basin relief above sea level could be used to predict sediment discharge. Although potentially problematic because basin area may be negatively correlated with basin relief, we calculate sediment discharge for a warm climate with a mean annual air temperature of 25°C in order to provide an alternative view of the potential power of rivers to erode Antarctica during the warm climate of the Eocene (Figure 8). This suggests that sediment discharge would range between 20 and 36 Mt/yr for the large radial drainage basins of East Antarctica, and below 10 for the short drainage basins controlled by the passive margin. These values are much smaller than the measured sediment discharges of 1,320 Mt/yr for the Amazon but are similar to the modest 58 Mt/yr for the Orange river at the present day (Summerfield & Hulton, 1994). Although speculative, the relief ratios and projected sediment discharge rates may provide context for understanding the relative contributions of these river systems to the sediments accumulating on the continental shelf surrounding Antarctica.

Analogy with present-day continental interiors gives an insight into the character of the land surface. The low river gradients imply low rates of erosion (Figure 8(b)). The calculated rate of denudation of the Lambert basin in pre-glacial times of 20 m per million years (Thomson et al., 2013) is similar to that of comparable southern continent rivers. Rates of

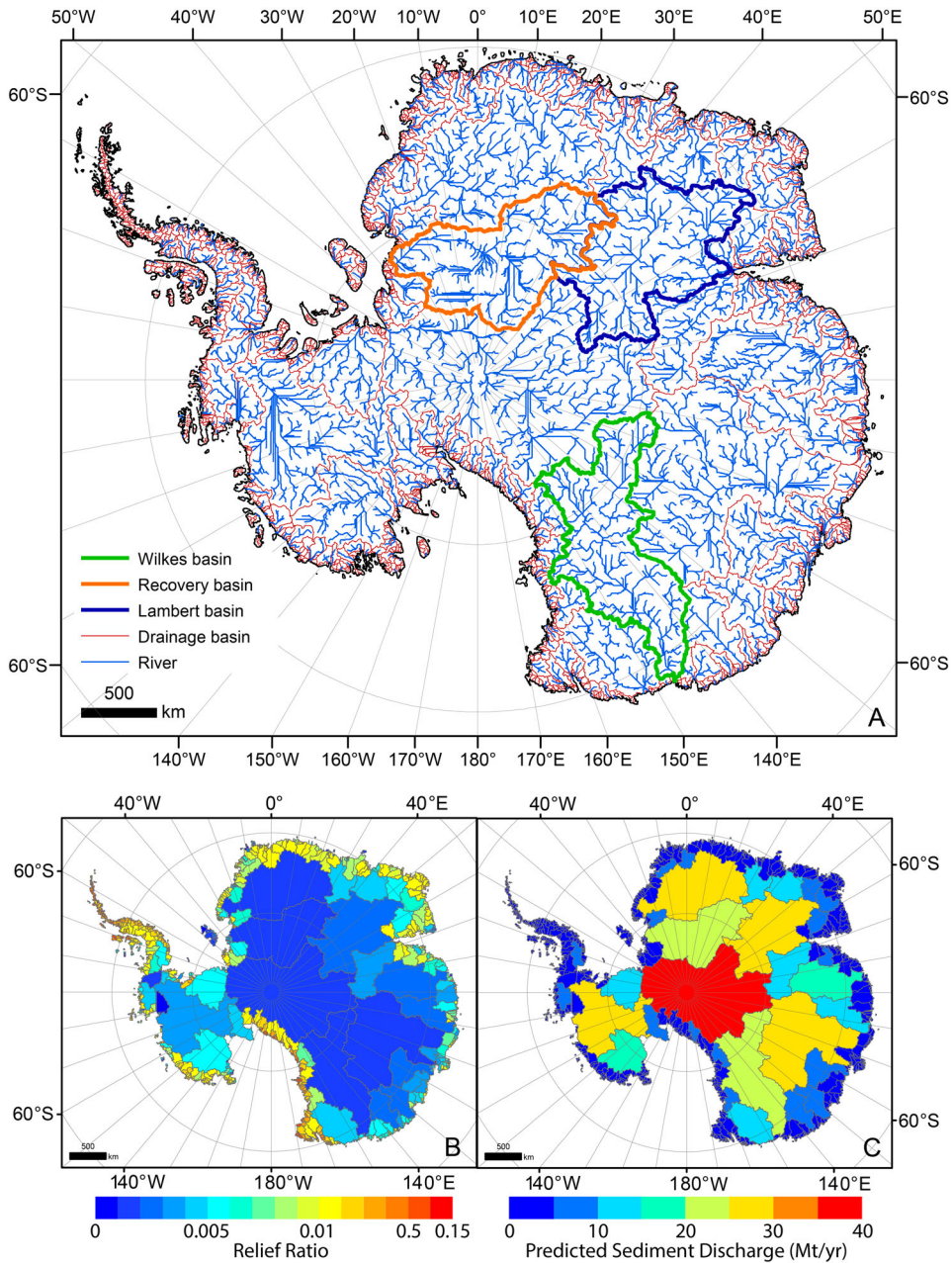


Figure 8. Pre-glacial rivers in Antarctica and estimates of denudation and sediment discharge. (A) Rivers in Antarctica as they would drain on the modern topography rebounded for ice loading. (B) Relief Ratio of Antarctic drainage basins as a proxy for denudation potential following Summerfield and Hulton (1994). (C) Calculated sediment discharge under a warm (25 °C MAAT) climate following Syvitski, Peckham, Hilberman, and Mulder (2003). The sediment discharge from the basin underlying the South Pole is 36 Mt/yr and only marginally more than several other basins. All these values are low in relation to the size of the basins and reflect low rates of denudation.

denudation of the La Plata, Zambezi, Orange and Murray basins are among the lowest in the world at 14, 15, 28 and 13 m per million years respectively (Summerfield & Hulton, 1994). Given the indications of a warm and wet Eocene climate for millions of years before glaciation we can infer deep weathering profiles such as in Madagascar and an abundance of corestones and inselbergs (Figure 5). Since the landscape is so ancient, the bornhardts could rise hundreds or even a thousand metres above the surrounding plain.

In West Antarctica are fault-bound micro-continentals that will also have experienced passive margin evolution. One would expect to find coastal lowlands with backing escarpments on those blocks adjacent to zones of crustal extension. One such example occurs in the Sarnoff Mountains of coastal Marie Byrd Land where dissected plateau remnants, some with tors and corestones, are surrounded by cliffs (Figure 9).

Two subglacial flat erosion surfaces separated by an escarpment have also been recorded on the Weddell Sea side of the Ellsworth block (Rose et al., 2015). Such features could well represent fluvial erosion of a passive margin. The presence of the highest mountains in Antarctica in the Ellsworth Mountain block is notable. The mountains rise abruptly above the eastern flank of a dissected plateau. The highest mountains, the Sentinel Range, consist of quartzites. With the exception of the plateau area around Mt. Vinson, they have a sharp crest and are bounded by a fault where they overlook Rutford Glacier that flows into the Weddell Sea. The valleys, now occupied by glaciers show a fluvial dendritic pattern (Rutford, 1972). Rock uplift of 4 km took place within 20 million years soon after break-up in the early Cretaceous (~140 Ma) and this was due to the isostatic unloading resulting from intense fluvial erosion of a similar thickness



Figure 9. Dissected plateau remnants of the Sarnoff Mountains, Marie Byrd Land, West Antarctica. Tors and weathered corestones occur on adjacent plateau remnants (Sugden, Balco, Cowdery, Stone, & Sass, 2005).

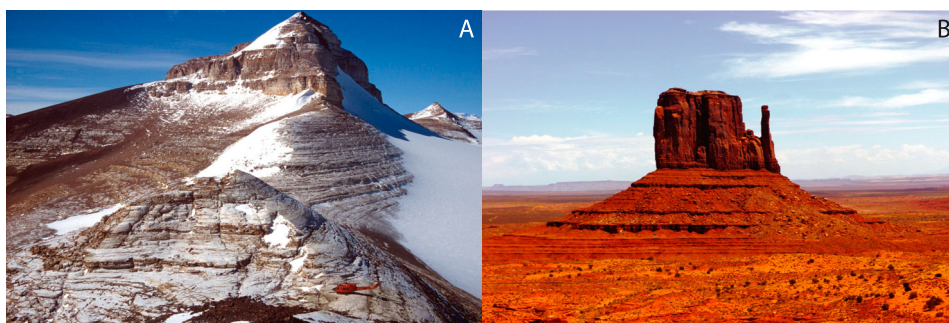


Figure 10. (A) Rectilinear slopes on the flanks of Altar Mountain, Quartermain Mountains, Dry Valleys. The thin regolith that covered slopes of 30° – 36° has been removed by overriding ice on the up-ice side to the right. (B) Similar rectilinear slopes in Monument Valley, Arizona, USA.

of rock (Fitzgerald & Stump, 1991). The combination of resistant lithology and proximity to a fault adjacent to crustal extension would have maximised topographic gradients and rates of fluvial erosion and rock uplift.

At a more local scale, the morphology of slopes may aid in the interpretation of landscape evolution. Straight or rectilinear slopes bounded by sharp breaks of slope are characteristic of the McMurdo Dry Valleys area (Figure 10). Indeed, Denton, Sugden, Marchant, Hall, and Wilch (1993) specifically note the similarity of the slopes of escarpments and inselbergs in the Olympus Range with the landscape of Monument Valley in Arizona. Straight slopes form when there is little vegetation and regolith and the main control is the angle of rest of weathered bedrock fragments. Surface wash removes any accumulation at the foot of a slope and evacuates the material across a low-gradient plain. Such conditions are characteristic of semi-arid climates. At present, it is difficult to reconcile this evidence of aridity with the vegetation evidence of humidity in Antarctica at least in the Eocene. Under the latter conditions one would expect convex and concave slopes typical of a landscape with regolith. One can speculate and suggest that aridity would have characterised the interior of Gondwana before significant break-up simply because of its great size. Perhaps the straight slopes relate to the pulse of erosion accompanying initial rifting before significant waterways open up between the continental fragments and led to a more humid climate. During early stages of rifting the river base level may have dropped while the climate remained semi-arid. As data on the Antarctic bed improves in resolution, it will be interesting to see the shape of slopes across the continent.

Wider implications for glaciation and passive margin evolution

The discussion above has tried to portray the Antarctic landscape before ice-sheet glaciation that commenced around 34 million years ago. It provides a tentative framework from which to improve understanding of the links between the fluvial landscape and ice-sheet stability. Elsewhere we have used an ice-sheet model to predict the glacial landscapes of Antarctica (Jamieson et al., 2010) and these are further quantified by Jamieson et al. (2014). The ice-sheet model predicts no erosion beneath cold-based ice and permits erosion beneath warm-based ice. Cold-based ice exists where the ice is thin and flow is slow or diverging, for example over mountains and uplands, and warm-based ice

extends over lowlands and locations where ice flow converges, particularly in former river valleys near the continental margins. Thus erosion is selective depending upon the pre-existing topography. The conclusion from the ice sheet-erosion modelling is that in areas of continuous warm-based flow in the interior erosion has removed less than 200 m in 34 million years (Jamieson et al., 2010). This figure is more than the tens of metres on shield areas subjected to northern hemisphere glaciation (Kleman & Hättestrand, 1999), and would be expected in view of the longer duration of Antarctic glaciation. A second conclusion from the modelling is that some glacial troughs, especially near the continental margin, were deepened by up to 2800 m.

Our reconstruction of the pre-glacial landscape of Antarctica adds perspective to the stability of the Antarctic Ice Sheet. A striking and relatively new observation is that huge troughs have been excavated into the lower reaches of the main river basins in East Antarctica. The trough within the Lambert Graben, up to 50 km wide, is cut selectively into the wider graben lowlands and extends 1000 km inland to the Gamburtsev Mountains. Both the ice-sheet model and study of offshore sediments suggest deepening of over 2 km (Jamieson et al., 2014; Thomson et al., 2013; Wilson et al., 2012). Within East Antarctica similar large troughs occur in George V Land, Queen Mary Land and an impressive concentration exists in Coats Land. In West Antarctica troughs with bases well below sea level occur between the main upland blocks and reflect ice flow into the Weddell Sea, the Ross Sea and the Pacific Ocean. In general, such troughs are larger than those cut by northern hemisphere ice sheets, again a feature that would be expected since they have been occupied by ice for much longer. These large Antarctic troughs have been called the weak underbelly of the Antarctic Ice Sheet since it is possible that further deepening might cause the ice to speed up, thin, retreat and thereby raise global sea level (Joughin & Alley, 2011). This process can be self-reinforcing if the glacier retreats into deeper water, for example into an over-deepened basin in the trough. Many scientists believe this is already happening in the case of Pine Island and Thwaites glaciers that flow into the Pacific Ocean between the Antarctic Peninsula and Marie Byrd Land (Rignot et al., 2014).

In possible contrast, there is a process whereby continued erosion by outlet glaciers flowing across an uplifted passive margin such as the Transantarctic and Dronning Maud Land mountains can increase ice-sheet stability. As the trough deepens, the unloading of the rock causes adjacent mountains to rise isostatically (Kerr & Huybrechts, 1999; Stern & Ten Brink, 1989). The higher the mountain rim the more stable the ice sheet might be because the isostatic response will cause the ice surface to be elevated into a colder climate, and because a barrier increasingly blocks ice flow from the interior. The importance of this factor will depend largely on the extent to which outlet glaciers in troughs crossing the mountain rim can maintain their flow and erosive capacity. Uplift due to isostatic unloading as ice replaces eroded rock amounts to hundreds of metres in the Transantarctic Mountains.

The pre-glacial landscape helps determine where glaciers first built up and the type of glacial landforms that resulted. Perusal of the landscape of the Gamburtsev Subglacial Mountains (Figure 7) demonstrates classic mountain glaciation features such as overdeepened and straightened troughs, wide valley floors, hanging valleys and corries (Bo et al., 2009; Rose et al., 2013). Such landscapes are ripe for more analysis. Mountains are likely to have played a significant role as glacial cycles waxed and waned. The highest mountains

will have held glaciers at an early stage of glaciation. For example, a glacial equilibrium altitude of 2000 m would be sufficient to create glaciers in seven mountain groups, namely Gamburtsev Subglacial Mountains, Antarctic Peninsula, Transantarctic Mountains, Marie Byrd Land, Ellsworth Land, Dronning Maud Land and Victoria Land (Jamieson et al., 2014). Viewed in this light it seems likely that mountain glaciers existed in cooler episodes in the Eocene before 34 million years ago. Indeed, it can be imagined that glaciers may have been juxtaposed with areas of temperate vegetation as can be observed in South America today. A threshold in ice-sheet growth would occur as glaciers left the mountains and extended over surrounding lowlands. Valley glaciers can flow relatively easily down a narrow valley, but as they reach the lowland they spread out into a piedmont lobe suddenly increasing the area of the ablation zone. In Chile and Italy, such valley glaciers repeatedly stalled at the mountain break and built large arcuate moraines. It will be interesting to see if such features are preserved and can be recognised around subglacial mountains in Antarctica.

The roughness of the ice sheet bed can influence the friction and thus flow and stability of an ice sheet. It seems reasonable to argue that a bed streamlined by glacial erosion will offer less resistance than a rougher bed. In this context ice-sheet modelling of glacial cycles reveals areas where ice erosion is in the same direction at all stages of a cycle and those areas where ice flow direction changes during a glacial cycle. The distribution of such areas has been predicted for Antarctica by Jamieson et al. (2010). Continuous streamlined flow in the same direction is most common where ice flow is in the same direction as the main river basins, for example in a broad zone from George V Land to Princess Elizabeth Land and in Enderby Land. But the interaction between ice flow from mountains of the central Transantarctic Mountains, Dronning Maud Land and the northwestern Gamburtsev Subglacial Mountains introduces complex changes in flow direction in interior East Antarctica during a glacial cycle. Perhaps such areas are rougher as a result. Perhaps, as in Scandinavia in zones where ice flow reversed during glaciation, they accumulate glacial debris from glacial cycle to glacial cycle (Kleman, Stroeven, & Lundqvist, 2008).

Offshore sediments can reveal much about rates and age of erosion. Reconstructing the pre-glacial landscape of Antarctica should help interpret the older sediments derived from passive margin evolution and also distinguish between the pre-glacial and glacial imprint. In turn, as shown in the case of the Lambert catchment, the analyses can help date landscape changes, for example, the timing of pulses of erosion and rates of erosion (Thomson et al., 2013). Offshore sediments also will help constrain the changes in both the regolith from smectite-dominated to illite-dominated clay minerals and in pollen in the sediments reflecting vegetation changes.

What is remarkable about the landscape of Antarctica is its apparent great age and the evidence of a landscape that has been evolving for tens and even a hundred million years. One reason for this is that the growth of ice sheets in the last 34 million years has protected the underlying landscape in zones where the basal ice is below the pressure melting point. Another reason may simply be that the cold climate of the last 14 million years means that running water has played a minimal role in weathering and transport of material. The latter helps explain how even minor features such as dolerite clasts, volcanic ashes and even buried ice have survived in the Dry Valleys for over ten million years (Marchant, Denton, & Swisher, 1993a; Margerison, Phillips, Stuart, & Sugden, 2005; Sugden, Marchant, et al., 1995).

Conclusion

Our focus on the evolution of passive margins under fluvial processes is an attempt to bring a coherent body of geomorphological thought to portray the landscape of Antarctica before glaciation. The overall conclusion is that the landscape is typical of those in other unglaciated southern continents and that the relationships between geophysical processes associated with plate tectonics and surface processes are the same. Since ice-sheet behaviour is influenced by interaction with the bed, the topic also has implications for ice-sheet stability. In the years ahead new, higher resolution bed topography datasets will become available and sampling of Antarctic bedrock geology will be undertaken. These will aid our understanding of the patterns and rates of landscape evolution. Our hope is that this paper provides a framework to help pose questions that can be answered from such data. It is also a challenge to geomorphology and physical geography worldwide!

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